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Using high-resolution stratigraphy to date fold and thrust activity: examples from the Neogene of south-central Sicily

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Abstract: The integration of structural and stratigraphic data is fundamental for determining rates of deformation in the uppermost continental crust. The high temporal resolution provided by Neogene marine sediments is used here to examine deformation rates in part of a thrust belt chosen from the Maghrebian orogen of Sicily. Conventional biozonal stratigraphy, calibrated against the geomagnetic polarity time scale, shows that individual thrust–fold structures grew steadily over many millions of years. Structures across the thrust belt were active at the same time and accommodated bulk shortening rates of *c.* 0.5 mm a^{-1}. In contrast the basal detachment operated about ten times faster. These results are in broad agreement with some theoretical models for orogenic wedge kinematics. Shore-line carbonate successions, calibrated with precession cycles of sea-level change, provide a very high-resolution, temporal scale with which to chart tilt rates on fold limbs (1°/27.6 ka = 0.036° ka^{-1}). These show that fold amplification was continuous although the rates may have varied with time. Incremental tilting of limbs during fold amplification is not predicted by popular models of fault-bend and 'fault-propagation' folding. Geometric modelling suggests that folding occurred by limb rotation, with minor hinge migration during buckling above buried thrusts. Thus stratigraphic data may be used to examine the kinematic evolution of thrust–fold systems along regional cross-sections, and of local structures. However, the types of structural models that can be tested using estimates of deformation rates depend upon the chrono-stratigraphic resolution available for the syn-tectonic sediments.

Keywords: Sicily, Italy, Neogene, thrust faults, folds, deformation rates.

Gaining reliable dates on the formation of geological structures is critically important for many aspects of the Earth sciences. Petroleum explorationists need to establish when anticlines formed relative to hydrocarbon migration (e.g. Demaison & Murris 1984). Within active zones of deformation, information on the timing and rates of structural evolution is essential in evaluating seismic hazard (e.g. Suppe *et al.* 1992). Comparisons of tilt and uplift rates from different parts of individual structures are needed to test kinematic and mechanical models of fold growth (e.g. Hardy & Poblet 1994). In thrust belts, information is needed on the relative timing and longevity of structures in understanding the evolution of orogenic wedges (e.g. Platt 1988). The challenge, then, is to develop dating strategies that provide sufficient resolution to meet these different needs. In this paper we examine the use of high-resolution stratigraphy to gain insight into deformation rates within a thrust belt. We review briefly the utility and limitations of the existing time scales before applying them to our study area within the Neogene thrust belt of central Sicily. Our initial motivation was to obtain a broad kinematic description of the thrust belt but this evolved into testing models of folding using the stratigraphy of synorogenic sediments.

For many years, interpretations of thrust belts have stressed sequential evolution, with faults moving in a specific order (e.g. Bally *et al.* 1966; Elliott & Johnson 1980), a view that continues today with an increasing recognition of so-called 'out-of-sequence' structures (e.g. Morley 1988; Mercier & Mansy 1995). Criteria for distinguishing thrust sequences generally rely on structural overprinting relationships (Butler 1987) which are well-developed when thrusts are closely spaced and consequently able to cross-cut each other or be folded by their hanging-wall structures. These settings are well displayed by the foothills-type structures of the Rocky Mountains (e.g. Dahlstrom 1970) and from more completely exhumed orogenic segments such as the Moine thrust belt (e.g. Elliott & Johnson 1980). However, many emergent structures are more widely spaced and consequently are not ammenable to overprinting criteria to establish relative age. Other, non-geometric, criteria are required.

Studies of aggradation rates within the Siwaliks of the frontal Himalayas of Pakistan (Johnson *et al.* 1986) suggest that several thrust-related folds may grow during the same period of geologic time along the same section line. For these examples it is inappropriate to describe structural evolution in terms of a strict sequence of thrusting. Corroboration that simultaneous thrusting occurs comes from consideration of fault-slip rates determined microstructurally, particularly from the Subalpine thrust belt of SE France. Here, arrays of slow-displacement thrusts, which are characterized by synkinematic calcite precipitation in shear fibres, must have acted simultaneously with fast-moving cataclastic faults to accomodate bulk shortening across the belt (Butler & Bowler 1995). Larger-scale geometric criteria for simultaneous thrusting have been described from the front ranges of the Canadian Rocky Mountains (Boyer 1992), where an imbricate fan in the Lewis thrust system was spawned from a thrust flat which is folded by coeval growth on a deeper culmination.

So the problem for at least some emergent thrust belts is not simply one of sequence but of activity. How long are thrust-related folds active, and at what rates? Is it possible to distinguish staccato-pulsed displacements and fold amplification from steadier, continuous activity? To address these questions, detailed information is needed on the longevity of folds, the rates at which they amplify and the rotation histories that their limbs have experienced. The most substantial information is provided by the stratal geometries of synorogenic sediments provided that they can be linked with some precision to a chronostratigraphic timescale and have identifiable stratal surfaces which record deformation.

Different degrees of stratigraphic resolution are required to address different tectonic questions. We start by using conventional biostratigraphic data to examine the large-scale evolution of structures seen in a cross-section. We then use a high-resolution stratigraphic technique based on depositional cyclicity to establish tectonic continuity.

Stratigraphic resolution: the importance of Neogene marine sediments

Few thrust belts preserve syn-orogenic sediments which were deposited across and adjacent to growing folds. Of those that do, much emphasis has been placed on subaerial anticlines where growth strata are recognised by syn-deformational unconformities and related stratigraphic onlap (e.g. Anadon *et al.* 1986; Williams 1993). Although stratal architectures may be understood in terms of syn-depositional folding, it is difficult in practice to use continental deposits as precise stratigraphic markers. In all but rapidly subsiding basins, such as the Himalayan foredeep of the Himalayas (Burbank *et al.* 1986), thrust anticlines are prone to erosion and so the critical stratal geometries necessary to reconstruct growth histories are progressively lost (Anadon *et al.* 1986). Even if preserved, it is very difficult to obtain a quantitative chronostratigraphy from continental sediments. They rarely contain useful biostratigraphy and consequently it is difficult to calibrate magnetostratigraphies, even if they are available. These problems may be avoided in examples where the syn-orogenic surface was submarine, where sediments could accumulate and be preserved during the growth of structures. There are additional advantages. Biozonal stratigraphy has a far higher resolution and there is far greater preservation of suitable fossils in most marine settings compared with subaerial cases. Palaeoecological data can be used to establish synorogenic bathymetries which, if combined with depositional analyses, can provide critical data on the evolution of slopes. The combination can provide constraints on time and geometry: this allows the determination of fold amplification rates with a greater precision than for subaerial structures.

Existing theoretical models of sedimentation at thrust anticlines (e.g. Suppe *et al.* 1992, 1997; Ford *et al.* 1997; Poblet *et al.* 1997) and many case histories (e.g. western Himalayas, Johnson *et al.* 1986) apply to simple patterns of sediment aggradation. In submarine settings these patterns will only be appropriate for the most distal parts, accumulating hemipelagic/pelagic sediments, coupled with very low tilt rates (compared with sedimentation). Elsewhere sedimentation will be strongly influenced by changes in relative sea level (uplift and subsidence together with eustatic fluctuations). Consequently depositional architecture must be analysed if the resolution of the stratigraphic record is to be used in full.

The highest stratigraphic resolution is provided by Neogene strata, particularly the Plio-Pleistocene (Berggren *et al.* 1995; Fig. 1). Within this period, a mixture of radiometric ages matched with astronomically driven depositional cycles in deepwater sediments has allowed calibration of magnetic chron and subchron boundaries with precisions of about 2–3 ka. This is achieved by defining a relative duration for

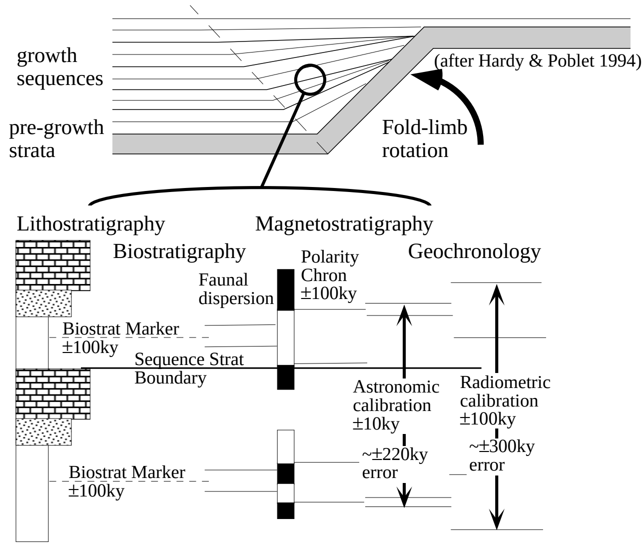

Fig. 1. The stratigraphic resolution available from growth sequences available for the determination of deformation rates. Lithostratigraphy alone may be of very limited use due to facies diachroneity. Conventional biostratigraphy for Neogene strata may have an uncertainty of ± 100 ka. Magneto-stratigraphy has a similar precision except when calibrated against astronomical cycles.

magnetostratigraphic chrons based on the width of sea-floor magnetic anomalies which may be calibrated against radiometric dates of basalts that retain the magnetic anomalies. Seafloor basalts are very difficult to date and the 181 reversals for the past 83 Ma are tied by only nine radiometric ages from terrestrial lavas (Cande & Kent 1992). However, for the Plio-Pleistocene, chron duration has been calibrated against depositional cycles in sediments which represent precession cycles in Earth orbit. The depositional cycles have average periodicities of 21 ka, with an accuracy of 1–2 ka. Thus each chron boundary for the Plio-Pleistocene that has been calibrated against astronomical cycles has an uncertainty of the duration of a single cycle. This can be more finely tuned using estimates of sediment accumulation rate, linked to palaeoenvironmental indicators such $\delta^{18}O$ cyclicity, to obtain precisions on chron boundaries of <5 ka (Berggren *et al.* 1995). Without this calibration, for older units, the uncertainties on chron boundaries are in the order of 100 ka.

The astro-magnetostratigraphy has been used to calibrate Plio-Pleistocene biostratigraphic datums (Berggren *et al.* 1995), particularly for nanno-plankton globally (23 in all) and, within the Mediterranean, planktonic foraminifera (58 in all). However, there are problems in the duration and diachroneity of fauna events (see Butler *et al.* 1996) which introduces substantial (1–2 Ma) errors in the absence of other corroborative stratigraphies. These are much greater for pre-Pliocene strata where there are, as yet, no calibrated high-resolution astronomical/geomagnetic polarity timescales. For Jurassic and Cretaceous fauna events, these uncertainties are substantially in excess of the average duration of the faunal events themselves (several million years; Harland *et al.* 1990). Consequently, late Neogene marine sediments offer the best stratigraphic resolution to obtain times and rates of geological processes.

The challenge of using high-precision cyclic sedimentation to chart fold growth is to use these sediments as structural markers. Fine-grained, deep-marine sediments are prone to slumping and commonly have few markers that can be traced with confidence across folds. In contrast, coarser-grained, shallow-marine sediments may have better stratal surfaces for charting deformation but do not contain the deep-water zonal fossils necessary for the accurate determination of ages. Later in this contribution we develop a correlation between deep- and shallow-marine stratigraphies that permits the marriage of suitable stratal geometries with a high-precision time-scale to chart fold growth.

Geological overview of the Sicilian study area

The Late Neogene basins in Sicily lie between the orogenic foreland represented by the Hyblean plateau in the SE ('northern Africa') and the metamorphic heart of the Maghrebian chain, represented by the Calabrian arc (e.g. Catalano & D'Argenio 1978; Lentini 1982; Fig. 2) in the NE. Sediments accumulated within foredeeps and thrust sheet top (perched) basins. The modern foredeep is the Fiumi Margi basin, within which has accumulated over 800 m of Quaternary clays (Butler *et al.* 1992). Compressional deformation has continued progressively so that early foredeep deposits, the Numidian flysch of Oligo-Miocene age (Catalano & D'Argenio 1978; Catalano *et al.* 1993), are incorporated in later structures and form the substrata to synorogenic perched basin fills.

In common with many circum-Mediterranean examples, much of the thrust belt developed in a submarine environment. Consequently sediments were able to accumulate across growing structures (Butler & Grasso 1993). Of the sequences deposited, those of Late Tortonian to Lower Pleistocene age are particularly well-preserved. The sequence stratigraphy and its tectonic controls, for Tortonian (Butler & Grasso 1993), Messinian (Butler *et al.* 1995*b*) and Pliocene (Butler *et al.* 1995*a*) sediments are documented elsewhere. In Tortonian times the region was generally submarine with substantial influxes of siliciclastic material derived from the orogenic interior. Syn-orogenic basin fills (Terravecchia formation) vary from proximal deltaic settings with high sediment delivery and accumulation rates to more 'distal' settings where pelagic–hemipelagic clays (Licata Formation) were deposited. In these basins regional tectonic subsidence (presumably flexurally driven from the orogenic belt) outpaced thrust-related uplift so that the tops of thrust anticlines could accumulate sediments. In Messinian times the Mediterranean experienced widespread desiccation. On Sicily this is represented by evaporite deposition. An angular, intra-Messinian unconformity, preserved in thrust-related synclines (Decima & Wezel 1973), is clear evidence for deformation continuing through this period. The return of fully marine conditions in earliest Pliocene times is marked by distal, deep marine chalks (the 'Trubi') and clays which 'shallow-upwards' into calcarenites. Foraminifera and macrofossils (e.g. Di Grande *et al.* 1976) evidence the bathymetric conditions and show that this sequence progrades from north to south across Sicily. The Plio-Pleistocene progradation developed as the foreland was unloaded by removal of the Calabrian arc by extensional faulting in the tectonic hinterland (e.g. Agate *et al.* 1993).

Regional cross-section and structural style

The south-central Sicilian fold and thrust belt is represented at outcrop by an array of widely spaced anticlines which bring up Tortonian and younger synorogenic deposits. Their sub-

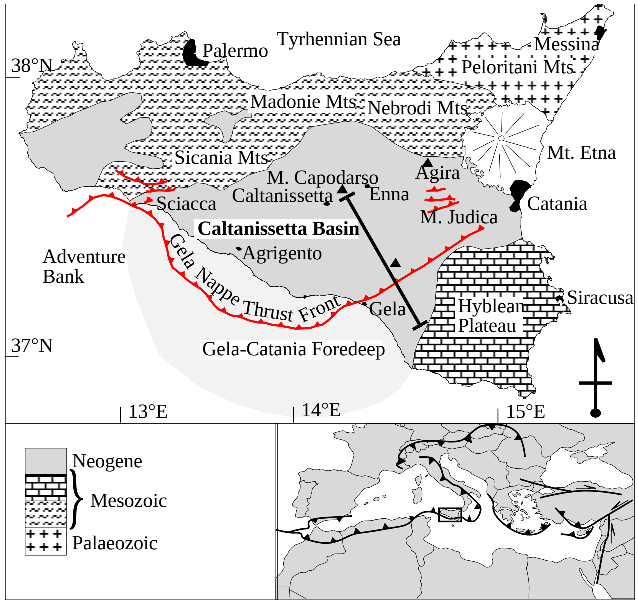

Fig. 2. Simplified map of Sicily (located on inset of Mediterranean). The line of section of Fig. 3 is indicated. The Fiumi Margi basin is the onland segment of the Foredeep NE of the town of Gela.

structure may be inferred from the Monte Judica and Sicania mountains along strike (Fig. 2), where thrusts carry Mesozoic carbonates onto Oligo-Miocene foredeep deposits (Carbone *et al.* 1990; Catalano *et al.* 1993). Consequently, we believe that the main fold belt overlies spaced imbricate thrusts (Fig. 3). This is confirmed by sparse drillhole data which penetrated through the allochthon into Pliocene chalks and the buried Hyblean foreland. The continuation of the underthrust foreland has been mapped geophysically by Grasso & Ben Avraham (1992).

The syn-deformational sediments are generally folded into periclinal structures with thrusts only rarely exposed at outcrop (Fig. 3), although several have been encountered in deep mines and wells. Thus it is likely that thrusts at depth terminate upwards into folds. The finite geometry is reminiscent of fault-propagation folding (e.g. Jamison 1987). However, there is little evidence of thrust surfaces growing during deformation as is required by such a description. It is likely that folding at shallow levels transfers down into thrusting at depth and in this sense the deformation is more akin to detachment folding, at least for the later stages of development. The relationships between faulting and folding is critical for understanding the kinematics of individual structures, a topic to which we will return later.

Large-scale stratigraphy: evidence for coeval thrusting

Tortonian and younger sediments show marked variations in thickness across thrust anticlines, with local depocentres in synclines (Fig. 3). It is possible for thickness variations to reflect sediment dispersal alone rather than tectonics with syncline growth preceeding the period represented by the basin fill. However, this can be recognized by considering facies and depositional environment of the sediments. For our examples, the Messinian marker (Calcare di Base) represents the same, marginal marine, palaeo-environment across the structures (Butler *et al.* 1995*b*) and, across our section-line, is broadly

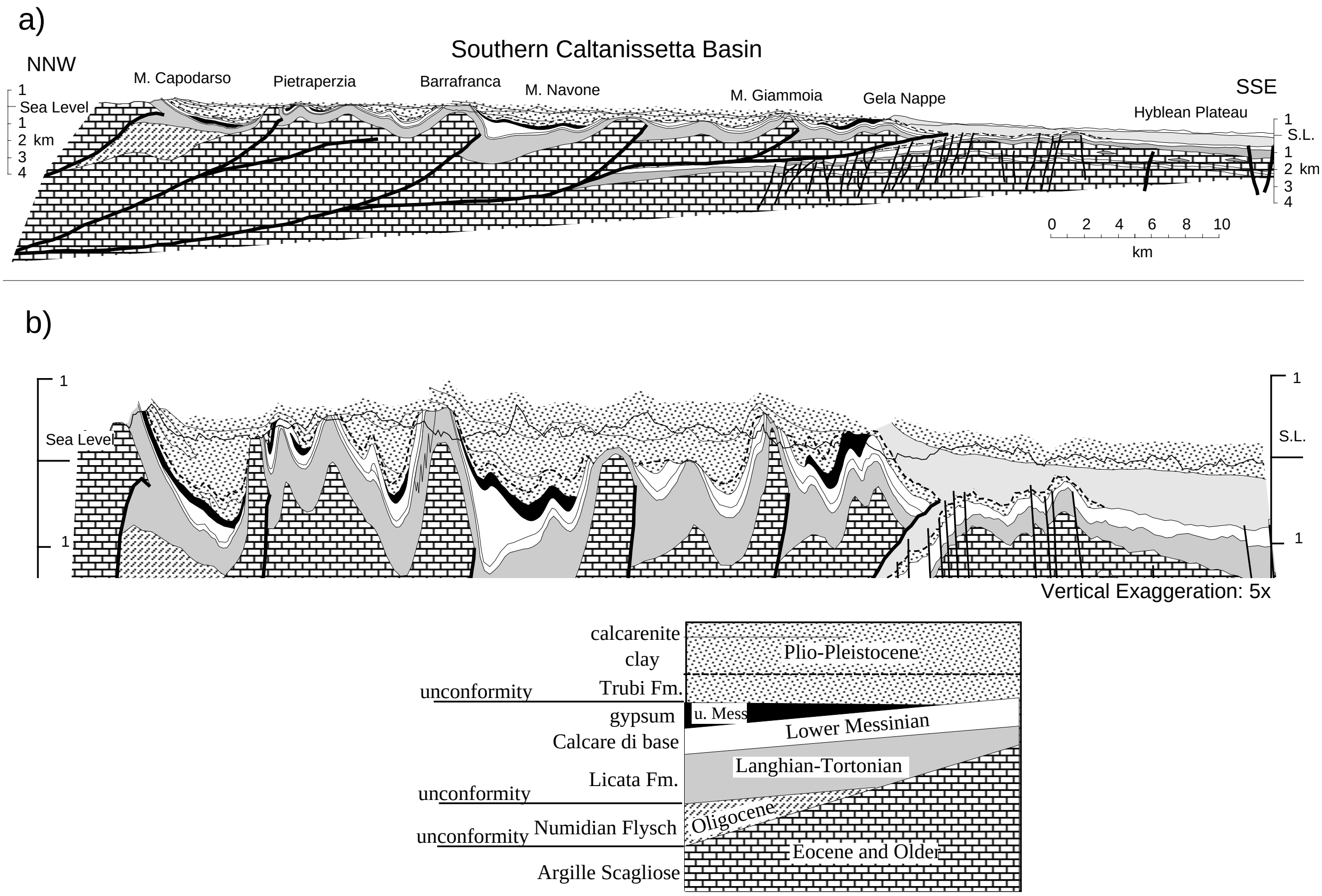

Fig. 3. Cross-section through the central Sicilian thrust belt (**a**) and a vertically exaggerated version (**b**) showing the geometry of syn-orogenic sediments collected in thrust-related synclines.

coeval (across a single chron boundary, McClelland *et al.* 1996). Consequently its cross-section shape represents post-depositional folding of an original, subhorizontal surface. The overlying Pliocene sediments are fine grained and prone to slumping. They onlap the Messinian on the flanks of structures, indicating that some folding of the Calcare di Base occurred prior to Trubi deposition. However, the Trubi is itself folded indicating that deformation occurred after its deposition too. The Plio-Pleistocene successions terminate upwards into shallow-marine carbonates and sands, capped by a regional subaerial surface. This forms an unfolded plateau indicating that deformation terminated with the deposition of the shallow marine sediments. These contain early Pleistocene fauna (Frazzetta 1971; Di Geronimo 1979; Butler *et al.* 1995*a*) indicating that deformation terminated regionally between about 1.6 and 1 Ma.

The tectonostratigraphic relationships described above are common to all the folds along our cross-section. Consequently we infer that all the folds were active during the late Miocene and Pliocene. We now investigate the incremental development of the cross-section by contrasting the amount of shortening experienced by different levels within the synorogenic sediments. The best constrained unit for measuring bulk strain is the Calcare di Base, a competent limestone that behaves as a 'key-bed'.

Using the Calcare di Base as a marker we calculate the total shortening during the late Messinian and Plio-Pleistocene. This unit is highly competent and so is amenable to simple line-length restoration. Its shape is projected where eroded across anticline crests. The cross-section was measured with an accuracy of 1 mm at a scale of 1:100 000 (100 m at true scale). The total length of the Calcare di Base from Monte Capodarso to the thrust front is 49.3 km over a present distance of 45.5 km. Therefore the shortening is estimated to be 3.8 ± 0.2 km. Regional magnetostratigraphic studies (McClelland *et al.* 1996) suggest that the top of the Calcare di Base broadly coincides with the base of the Gilbert, Chron C3r, dated at 5.89 Ma (Cande & Kent 1995). Thus the total duration of deformation was from 5.9 Ma to between 1.6 and 1 Ma, a duration of 4.6 ± 0.3 Ma. So the shortening rate is estimated as being 0.83 ± 0.1 mm a^{-1}.

A similar calculation may be performed for the top of the Trubi Formation but this facies boundary is diachronous by perhaps as much as 500 ka across the cross-section (Butler *et al.* 1995*a*). Furthermore, the Trubi is incompetent and can contain a weak cleavage. Consequently the shortening estimates obtained by line-length measurement are prone to an unquantifiable error. We assume that they slightly underestimate the true shortening. The arc-length of the top of the Trubi is 47.0 km over the present-day distance of 45.5 km, giving a shortening value of 1.5 ± 0.1 km. If we place the age of the top of the Trubi as 3.25 ± 0.25 Ma (Butler *et al.* 1995*a*), the duration of deformation was about 1.9 ± 0.5 Ma. Thus the shortening rate was 0.86 ± 0.28 mm a^{-1}.

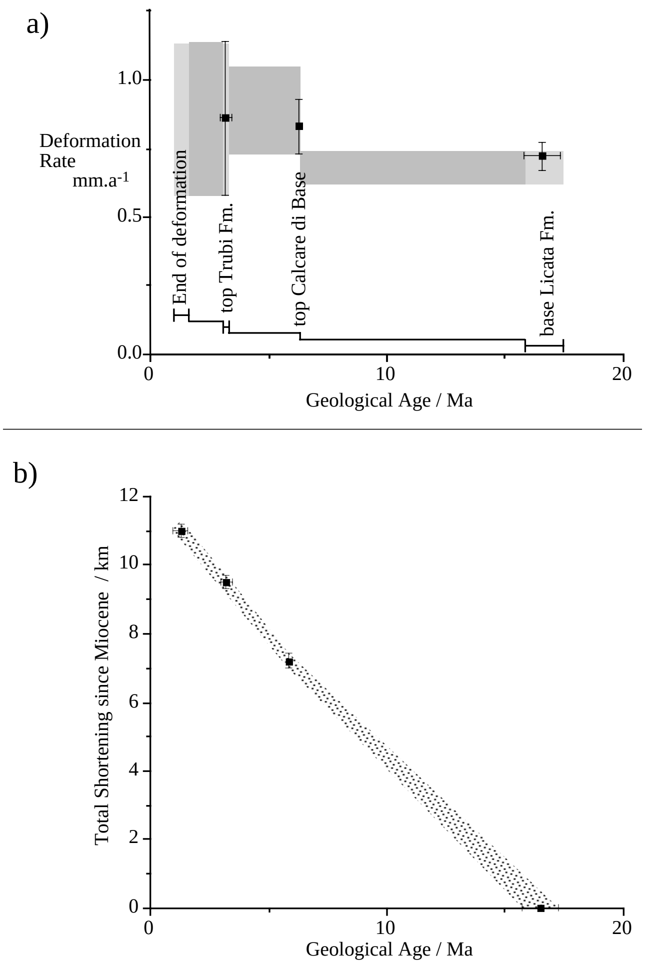

Fig. 4. Variations in deformation rate (**a**) and cumulative shortening (**b**) through time for the cross-section of Fig. 3.

Finally we can use the base of the Licata/Terravecchia formations, at the base of the thrust-top sequences. The cross-section (Fig. 3) shows the arc-length of this level to be 56.5 km, with the same present distance of 45.5 km. So the shortening is 11.0 km. Although it is possible to obtain this measurement from the cross-section itself, it is not very well constrained at depth and so the value above has a significant, but unquantifiable error. However, it is likely to be near a minimum estimate, not only because of extra, buried bed-length in the footwall to thrusts but also because of distributed layer-parallel shortening strains due to the formations being fine-grained and, like the Pliocene units, prone to volume changes. The formations contain Langhian fauna, placing the base of the unit at 16.5 ± 0.8 Ma. So the total timespan over which 11 km of shortening was accumulated is 15.2 ± 1.1 Ma. The time-averaged shortening rate was 0.72 ± 0.05 mm a^{-1}.

The rates of shortening can be compared for each time-increment, working back from the youngest (Fig. 4). The last increment, as recorded by the top of the Trubi alone (the final 1.5 ± 0.1 km shortening), is reported above as 0.86 ± 0.28 mm a^{-1}. The Calcare di Base had accumulated a total of 2.3 ± 0.2 km shortening (3.8 ± 0.1 minus 1.5 ± 0.1 km) prior to deposition of the last part of the Trubi. The period between the end Calcare di Base (5.89 Ma) and the top of the Trubi (3.25 ± 0.25) is 2.64 ± 0.25 Ma. So the shortening rate for this increment was 0.89 ± 0.16 mm a^{-1}. The base of the Licata/Terravecchia formation had accumulated 7.2 ± 0.1 km shortening prior to the end of Calcare di Base deposition. This period had a duration of 10.6 ± 0.8 Ma so the shortening rate for this first increment was 0.68 ± 0.06 mm a^{-1}. The errors on these figures that are unquantifiable, essentially relating to the distributed strains in the dominantly fine-grained sediments, render the apparent precisions and variations on the above numbers difficult to compare. Nevertheless, it is likely that the bulk shortening rate was slow throughout the evolution of the structures and operated at less than 1 mm a^{-1}.

Displacements rates on the basal detachment

The shortening estimates developed above only concern deformation within the thrust belt. The basal detachment is excluded from the estimate. The allochthon overthrusts chalks of the Trubi Formation by at least 6.5 km. The chalks are younger than the base of chron C3n.4n (Langereis & Hilgen 1991), dated at 5.23 Ma (Cande & Kent 1995). Displacements continued until early Pleistocene times (1.6–1 Ma). The maximum duration for thrusting must have been 4.23 Ma at a time-averaged rate 1.5 mm a^{-1}. However, if the bulk of the displacement occurred after the end of Trubi deposition (about 3 Ma, Butler *et al.* 1995*a*) the time-averaged displacement rate on the basal thrust was 3.25–4.6 mm a^{-1}. This value is a minimum estimate because the displacement magnitude on the basal detachment could be substantially greater than is constrained by available well-holes that penetrate the buried foreland (Butler *et al.* 1992).

Deformation partitioning

The above analysis shows that deformation in the south-central Sicilian fold and thrust belt is partitioned between the basal detachment and the thrust belt itself. Individual thrusts appear to be active for many millions of years at individually slow shortening rates. The basal detachment was certainly active at much faster rates for the period after the deposition of the Trubi (late Pliocene, early Pleistocene) than the internal thrusts. Within our time frame, thrusts were active simultaneously and in parallel. However, the stratigraphic resolution used above is insufficient to distinguish between continuous thrust and fold growth or more staccato deformation. Further resolution is theoretically possible using Plio-Pleistocene biozones but there are practical difficulties tracing these in monotonous pelagic sediments in the field.

High resolution stratigraphy to investigate fold kinematics

Arguments about the detailed amplification history of folds require a degree of stratigraphic resolution significantly higher than that obtained by even magneto-stratigraphically calibrated faunal zonations. However, the astronomical cycle stratigraphy that is well-calibrated for Plio-Pleistocene times in the Mediterranean (reviewed by Berggren *et al.* 1995) provides a possible tool with time increments of about 23 ka.

Ideally, the growth of folds should be studied using measurements of relative uplift of a single surface across the fold profile (Fig. 5). This approach requires a recognizable reference level from which differential vertical movements can be measured. For syn-tectonic sediments this reference is the palaeo-Earth surface. Practical difficulties arise in using many

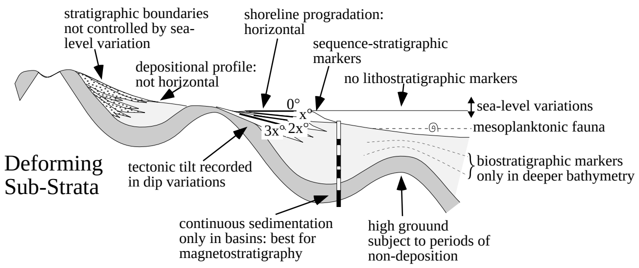

Fig. 5. Hypothetical illustration of the use of syn-orogenic strata to determine uplift and tilting rates across active folds. Different palaeo-environments offer different compomises between determining palaeo-elevations/slopes, preservation potential and chronostratigraphic resolution.

parts of the earth's surface because its level can vary because of non-tectonic processes, particularly sedimentation and erosion. Consequently a reference level is needed for the land surface: in practice this is sea-level. Across an area of interest, sea level will have varied not only due to the locate amplification of the fold and fault structures which are being analysed but also due to glacio-eustatic fluctuations and regional tectonics. Nevertheless, sea level is a useful marker, provided the entire fold structure has experienced the same base level evolution and that time-lines, at the appropriate stratigraphic resolution, can be correlated across the structure.

One approach is to choose examples with uniform sediment supply so that syn-orogenic strata experience differential aggradation across an amplifying fold (Fig. 5). The fold generally has no expression at the depositional surface because it is entirely buried under a blanket of sediment (Poblet *et al.* 1997). Although ideal differential aggradation underpins many semi-quantitative models of syn-tectonic deposition (e.g. Suppe *et al.* 1992; Hardy & Poblet 1994), it is difficult to test in many real examples. The best settings are provided by depositional environments situated at or very close to sea level. Unfortunately, coastal plain and shallow-marine sediments are difficult to date. The zonal fossils calibrated by Berggren *et al.* (1995) were free-swimming: the principal Plio-Pleistocene biozone foraminifera (the Globorotalids) were mesoplanktonic (Cita 1975) and lived in water depths of >50 m. For these depths, below storm wavebase, it is difficult to obtain precise palaeobathymetric estimates. Thus for good biostratigraphic control, essential to obtain useful estimates of rates, it is difficult to also get estimates of differential vertical movements with errors of less than perhaps 100 m. The degree of uncertainty greatly limits the application of these type of stratigraphic data in fold studies.

An alternative approach is to study the evolution of part of the fold geometry, specifically the limb-tilt rates (Fig. 5). Incremental limb dips can be measured provided that accurate estimates of the syndepositional slope can be made. Again, this limits the number of suitable palaeoenvironments. Coarse grained fans, commonly found adjacent to actively eroding folds, are probably the least-suitable materials because of their high primary slopes and the uncertainties in assessing them. Here we outline the utility of coastal calcarenites from the microtidal or atidal Mediterranean, a palaeo-environment that was very sensitive to variations in tilt and baselevel changes yet can also be correlated with zonal fauna.

A detailed depositional and sequence stratigraphic model for the calcarenites is presented elsewhere (Lickorish & Butler 1996). Found as 5–10 m thick packages with internal clinoform

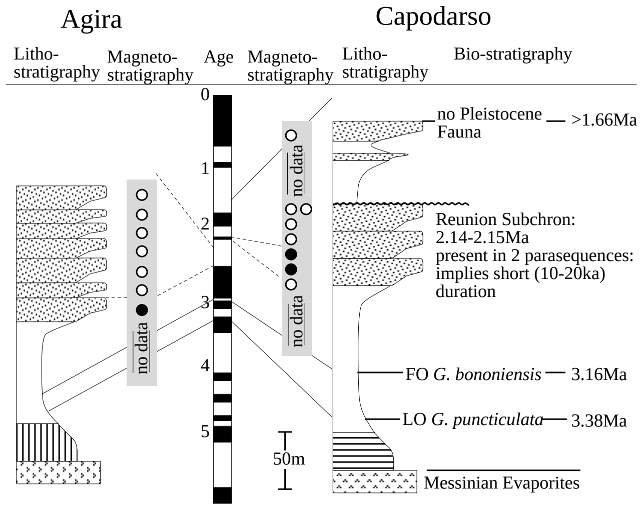

Fig. 6. Chronostratigraphic calibration of Pliocene calcarenites, organized within cyclic parasequences, for the Agira and Capodarso sites. Further geological details are provided by Butler *et al.* (1995*b*) and Lickorish & Butler (1996).

architectures, the sediments are packstones composed of fragmented shelly fauna. The sands represent the top of shoaling-upwards units, repeated cyclically. Each cycle is interpreted as a parasequence and terminates upwards in facies that represent normal wave-base (1–4 m bathymetry). Because the Mediterranean was atidal during the late Cenozoic, the parasequence tops represent simple surfaces that had a syn-depositional slope <1° and a palaeobathymetry of <5 m. Consequently they are excellent palaeo-environmental markers to chart dip and elevation changes on folds.

The regular cyclicity of Plio-Pleistocene shallow-marine calcarenites and near-shore muds strongly suggests a eustatic control, with stacking patterns modulated by tectonics (Lickorish & Butler 1996). To use these architectures to derive tectonic rates demands placing the parasequences into an appropriate order of eustatic cyclicity. We do this first and then go on to document tilt rates on folds.

Parasequence durations

We have calibrated the duration of each parasequence, using a type section at Monte Capodarso (Fig. 6). Here there are six calcarenite units. Because of the high-order eustatic signal in the late Cenozoic, the calcarenites are interbedded with deeper marine sediments that contain zonal foraminifera (Fig. 6). The depositional cycles of correlate well with the highest order (precession) cycles in eustasy (Lickorish & Butler 1996). Precession cyclicity has been calibrated against the geo-magnetic time scale for the Plio-Pleistocene using deep water sediments in the Mediterranean (Hilgen 1991; Berggren *et al.* 1995). Each cycle represents 23 ka (±2 ka) and so we infer that each parasequence is of similar duration. Clearly each calcarenite unit represents just a small fraction of this.

We have tested our calibration using magnetostratigraphy (Lickorish & Butler 1996). Although the parasequence stack is dominantly represented by reversed polarities, two units record normal magnetisations. The biostratigraphy places the whole parasequence stack to be within the Matuyama, C2r chron, with the small segment of normal polarity representing the Réunion subchron (Fig. 6). This subchron is calibrated as

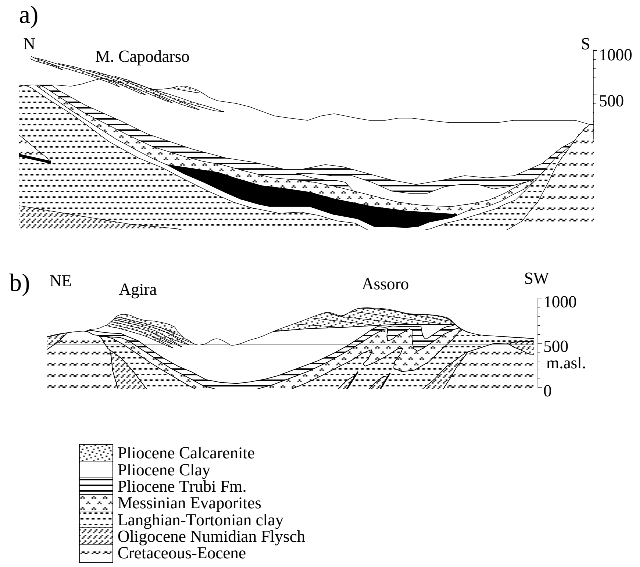

Fig. 7. Cross-sections through the M Capodarso (**a**) and Agira (**b**) sections (no vertical exaggeration) showing the architecture of Pliocene calcarenites. At Capodarso the parasequences are organised into an oflapping array while at Agira they form an aggradational stack. After Butler *et al.* (1995*a*).

between 2.14 and 2.15 Ma (Berggren *et al.* 1995). These data support our interpretation that the parasequences record precession and that each represents about 23 ka.

Controls on parasequence stacking patterns: local versus regional tectonics

Although parasequences record high-frequency eustatic variations in sea level, this record is severely modulated by tectonics. Tectonic controls are exerted far-field, through regional movements associated with flexural support of the foredeep basin and locally through differential uplift at thrust-related anticlines as discussed above. These two tectonic controls can be separated by consideration of the parasequence stacking patterns.

Aggradational parasequence stacking requires renewal of accommodation space across several eustatic cycles. This is most plausibly achieved by tectonic subsidence. For aggradation across the crests of anticlines subsidence is required to outpace the uplift associated with fold amplification. In contrast, fold amplification may result in progressive limb rotation and this will be represented by systematic reductions in dip upsection for the parasequence boundaries.

These two behaviours are well-displayed by the growth fold at Agira (Fig. 7; Butler *et al.* 1995*a*). Here seven parasequences have been identified. Six are stacked above each other on the crest of the fold. On the southern fold limb there is a systematic decrease in the dip of parasequence tops upsection. Linked bio- and magneto-stratigraphy show these mixed behaviours on anticlines up to about 2 Ma (Butler *et al.* 1995*a*). Younger folds are characterized by systematic offlap without aggradation at fold crests. This indicates that foredeep subsidence continued until about 2 Ma and thereafter the region has rebounded. The regional regression rate is inferred to have rapidly increased at this time.

Limb-tilt rates and fold models

Plio-Pleistocene calcarenites dominate the geology of south-central Sicily and a number of excellent sites for study of these sediments at growth folds are available. Generally, sites show basin sections in the synclines dominated by muds and clays with a rich foraminifera fauna. The fold limbs contain divergent sheets of calcacarenites with internal clinoforms which we interpret as stacks of parasequences. Although the Agira section, discussed briefly above, shows well the growth and tilt evolution in the parasequences, the underlying structure of the older strata is only poorly known, particularly in the critical area beneath the northern anticline (Fig. 7). Consequently we now discuss a site at Monte Capodarso (Fig. 7a) which is better understood because of a vast subsurface dataset from salt mines.

Tilt rates at Capodarso

Monte Capodarso lies on the southern flank of the Marcasita anticline (Fig. 3), a thrust-related fold that was active through the Mio-Pliocene (Fig. 6; Butler *et al.* 1995*b*). The syn-orogenic deposits are capped by the Capodarso calcarenites (Fig. 7), which are ordered as an offlapping array of six parasequences (Butler *et al.* 1995*a*). The upper surface of the youngest parasequence dips at 22° while the oldest dips at 27°. Our calibration of the duration of each parasequence (23 ka) implies that the stack represents 138 ka during which time 5° of tilting occurred. Thus the tilt rate for the period represented by the parasequences was $1°/27.6\ \text{ka} = 0.036°\ \text{ka}^{-1}$.

The evidence from the calcarenite sequence provides a high-resolution estimate of the tilt rate over a short period of time, and suggests that there is a continuous, slow rotation of the fold limbs over time. The longer term tilt history of the fold limb can be examined by adding information from older syn-orogenic layers. The Pliocene calcarenite succession (2.1 Ma, dipping 22°) and the Messinian Calcare di Base (5.89 Ma, dipping 40°) provide two control points across the structure where both the dip and the absolute age of the strata are known with some precision, separated by rheologically weak clays within which the stratigraphy is poorly constrained. Using the previously described values for the stratigraphic ages the duration of deformation is 0.8 ± 0.3 Ma after the calcarenite and a further 3.79 Ma back to the Calcare di Base, indicating a progressive tilting for over 4 Ma. This yields an average tilt rate of $1°/114.8 \pm 7.8$ ka ($0.0088 \pm 0.0006°\ \text{ka}^{-1}$), which can be divided into $1°/21.6$ ka ($0.0048°\ \text{ka}^{-1}$) for the time between the Calcare di Base and the calcarenites, and $1°/36.3 \pm 13.6$ ka ($0.032 \pm 0.012°\ \text{ka}^{-1}$) for the time span after the calcarenites.

Comparisons with theoretical folding models

There have been several published attempts to use stratal architectures of syn-orogenic deposits to establish rates of fold growth (e.g. Suppe *et al.* 1992). These generally use finite stratal geometries where a simple kink-band fault-bend or fault propagation fold model is assumed (Suppe *et al.* 1997). The validity of applying this type of model can be tested directly using stratal architectures. For this purpose we examine three folding models (Fig. 8).

Ideal kink-band type models (Suppe 1983) require the migration of fold axial surfaces through strata during fold amplification (Fig. 8a). An individual bed-segment experiences

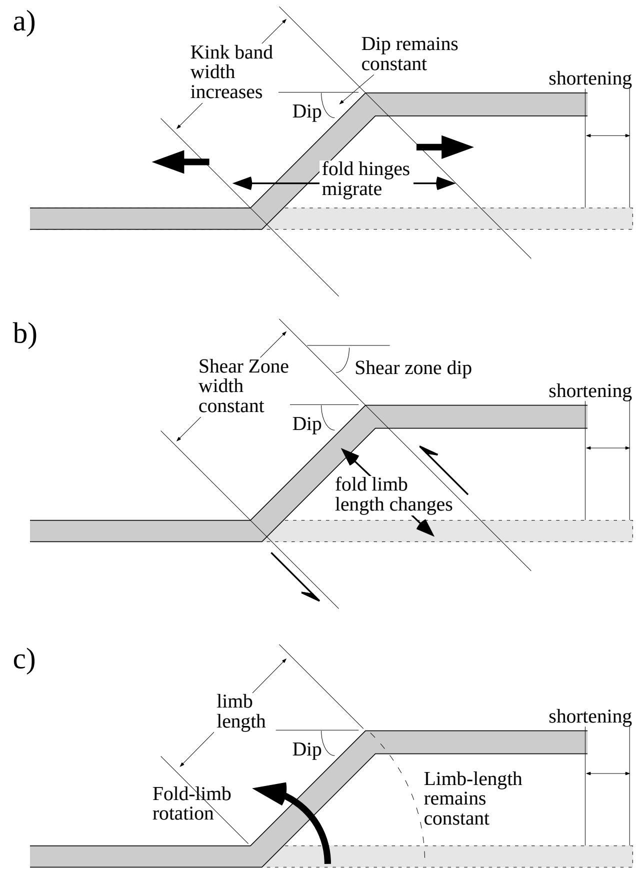

Fig. 8. Kinematic models for fold growth. (**a**) Constant limb-dip, hinge-migration theory. This model is inapplicable to examples that show incremental tilt histories in growth strata. (**b**) Limb rotation in a shear zone. The fold limb strains to vary bed-length but the fold hinges are fixed. (**c**) Simple limb-rotation with fixed hinges. This model does not balance (Butler 1992) and requires minor hinge migration with limb rotation during fold amplification.

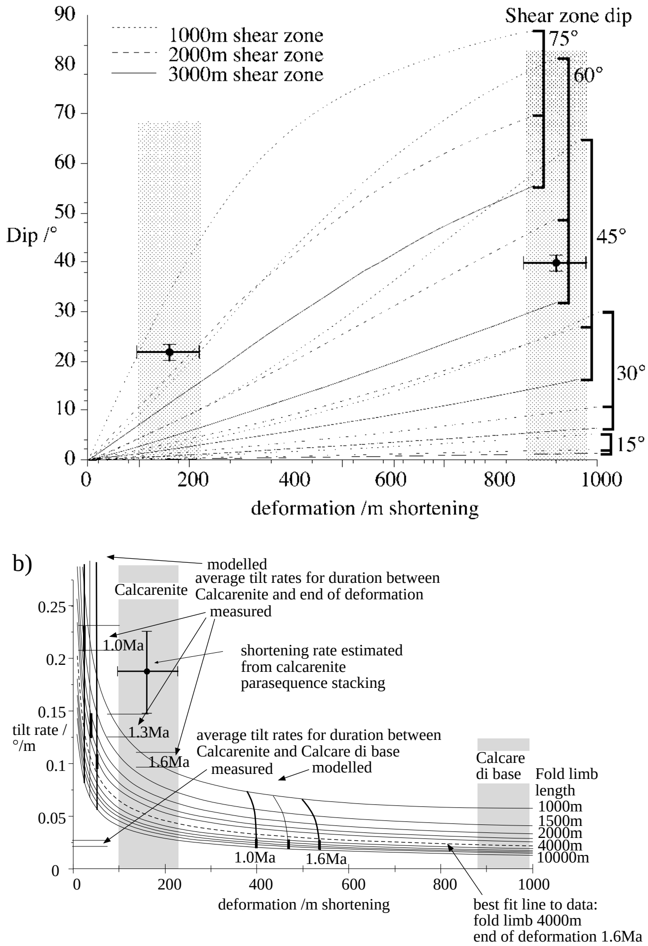

Fig. 9. Curves showing the theoretical relationship between bulk shortening and limb dip for the shear zone model (Fig. 9b). Data are overlaid for the upper calcarenite and the Calcare di Base. See text for further discussion.

an instantaneous rotation as the axial surface sweeps through. Clearly such a model is inconsistent with a progressive limb rotation such as is recorded at Capodarso.

A second kinematic folding model involves progressive limb rotation as part of a shear zone effectively bounded by the two fold axial surfaces (Fig. 8b). In this model the width (w) and dip (φ) of the shear zone are kept fixed and these two parameters control the relationship between the amount of tectonic shortening (s) and fold limb dip (θ) such that:

$$\tan \varphi = s\tan \theta/(w\sec \theta - s).$$

The results of this are plotted in Fig. 9, illustrating the differences for a variety of values of both shear zone dip and shear zone width. The model is constructed independent of time and deformation rates. However our field data points are calibrated with respect to time and not tectonic shortening. Estimates of the tectonic shortening must therefore be calibrated in order to make a comparison. The average tectonic shortening rate is constrained by the total shortening of the Calcare di Base across the structure (*c.* 1 km in 4.89 Ma) and can be cross-checked by the estimated regional shortening rate and the number of active major structures (*c.* 0.8 mm a^{-1} across four major anticlines; Figs 3 and 4). Since there is no regional evidence for significant variations in tectonic shortening rates over this period, a constant value of 0.2 mm a^{-1} is assummed. The data points can then be plotted on Fig. 9 by calculating the total shortening the fold experienced after deposition of the marker. The error bars represent the uncertainty in the age of the end of deformation. The actual value used is not that critical since most of the comparisons are qualitative, but a realistic value generates reasonable predictions.

The general behaviour of the Marcasita fold limb at Monte Capodarso is explained where early, syn-deformational strata are more steeply dipping and have experienced greater rotations than the younger strata. However, assuming that the shear zone dip, shear zone width and tectonic shortening rate remained constant with time, these two data should lie on the same theoretical curve. For all reasonable values of these parameters they do not, since the theoretical curves are concave in this region, and the empirical curve through the origin is convex. Thus either the parameters did not remain constant with time or the kinematic model is inapplicable. We prefer the later explanation, for an additional, qualitative reason. The model predicts significant longitudinal strains on bedding planes during limb rotation and there is no record of such strains in outcrop.

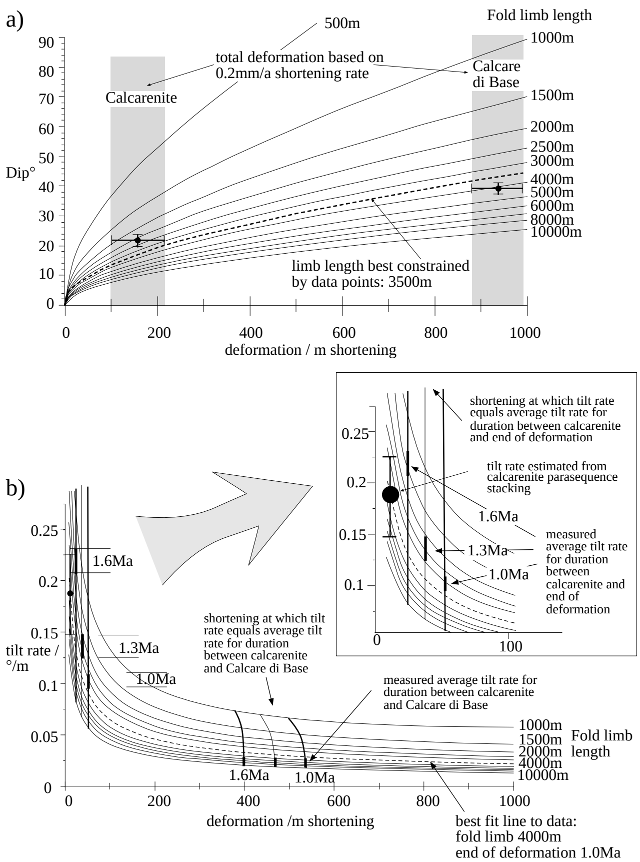

Fig. 10. (**a**) The relationship between bulk shortening and limb dip for the fold model of Fig. 9c. The curves show the predicted dip of a bed after it has undergone a specified shortening. (**b**) The curves are the derivatives of those in part (a) showing the tilt rate of a bed after it has undergone a specified amount of shortening. Also shown are curves for the shortening at which the instantaneous tilt rate equals the average tilt rate over the two time-periods from the end of deformation to the calcarenites and from the calcaernites to the Calcare di Base. Three lines are plotted for each corresponding to the minimum average and maximum estimates of the age of the end of deformation. The range of the empirically derived values are plotted as heavy line segments. The instantaneous tilt-rate recorded during the calcarenite deposition is also plotted.

A third kinematic model has limb rotation but with limb-length remaining constant (Fig. 8c). As with the shear zone model (Fig. 8b), the fold hinge remains fixed with respect to bedding. This rotational kink-band model does not require longitudinal strains on bed-segments. In this case the relationship between fold limb dip (φ) and shortening (*s*) is only dependent upon limb length (*w*) such that:

$$\cos \varphi = 1 - s/w.$$

The results of this are plotted in Fig. 10a for a variety of values of the limb length. In all cases the curve is convex such that early increments of dip accumulate more rapidly than the later ones, an attribute consistent with the field data.

The best-fit limb length for the data from the theoretical curves is 3500 m, a value consistent with the visible structure (Fig. 7). For a further test of the model the theoretical evolution of tilt rate with the increment shortening (Fig. 10b, derived from Fig. 10a) can be compared with various geological estimates for Capodarso. Taking various ages for the end of deformation in the section (1.0 Ma, 1.3 Ma, 1.6 Ma) the theoretical shortening is calculated at which the tilt rate would equal the average tilt rate over the time intervals represented by the two time periods between the end of deformation, the calcarenite deposition and the Calcare di Base. Still using the estimated shortening rate of 0.2 mm a^{-1}, the average tilting of the Calcare di Base for the period from its deposition to the first calcarenite parasequence (0.0048° ka^{-1}) is just 0.024° per metre of shortening, this low value of tilt rate implies that a very long fold-limb is needed. However, values for the average tilt rates of the calcarenite imply a shorter fold limb length. A curve with a limb length of 4000 m just about fits within the margins of error if the 1.0 Ma age for the end of deformation is taken. The final piece of data on Fig. 10b is the instantaneous tilt rate of the calcarenite parasequences during deposition (0.036° ka^{-1} equivalent to 0.18° per metre of shortening) which is a much higher value than any of the average values. However the differential tilting within the calcarenite sequence represents deformation within the first 138 ka after deposition of the first parasequence and so represents about the first 30 m of tectonic shortening for which very high tilt rates of this magnitude are predicted. This model of a fold with a rotating limb of fixed limb length seems to fit the observed data quite well without the need to invoke variations in tectonic shortening rate.

By appealing to variations in shortening rates through time any values of fold limb tilting can be generated. The data presented here are shown to be compatible with a model that requires no changes in bulk tectonic shortening rate over a time period of at least about 4 Ma.

Conclusions

At present, stratigraphic data represent the most important records of geological time that can be used to date structures. The resolution provided by combined magneto- and bio-stratigraphies for the Neogene allows fault and fold activity within evolving orogenic wedges to be assessed. Our case history from Sicily shows that different thrust structures within the orogenic wedge were active simultaneously, as predicted by Platt's (1988) theoretical model of imbrication within orogenic wedges. The shortening rate across the thrust belt was about 0.5 mm a^{-1}, with slow, continuous amplification of individual thrust anticlines. These structures collectively represent the products of continuous horizontal shortening within the orogenic wedge. The stratigraphic and structural resolution is less for the basal detachment of the orogenic wedge than for the imbricate thrust within. However, it is sufficient to establish minimum time-averaged slip rates of 3.25 mm a^{-1} for the later part of its history. These values pertain for the last few milion years of structural evolution. If they are generally representative of orogenic wedges, it implies that the active basal detachment accommodates approximately ten times more shortening than internal deformation within the over-riding thrust belt. If fault slip rates reflect fault zone rheology, it is likely that the mechanics of faulting on these different types of thrust are very different.

Within the resolution provided by traditional, bio-zone-based, stratigraphy, the individual thrust anticlines appear to have amplified continuous for many millions of years. When sedimentation is continuous (over the 100 ka timescale), there are no abrupt angular discordances as would occur if fold amplification was jerky (e.g. Anadon *et al.* 1986).

The high degree of temporal resolution possible from identifying astronomically controlled sedimentary cycles permits the investigation of fold kinematics. However, the stratigraphic resolution must be matched by the ability to constrain the depositional geometry and elevation of stratal surfaces across folds. By studying Pliocene coastal calcarenites in central Sicily we have been able to satisfy both requirements. These particular calcarenites are organised in parasequences which calibrate with precession cyclicity of *c.* 23 ka periodicity. Their top surfaces represent palaeo-horizons ($\pm 1°$) at a well-constrained palaeobathymetry (<5 m). Tilt rates of 1°/27 ka (0.036° ka^{-1}) for the first *c.* 130 ka of the history of tilting these beds, although long term averaged rates are considerably lower e.g. 1°/114.8 ka (0.0088° ka^{-1}) for the total evolution of the Calcare di Base.

The tilt data may be used to test kinematic models for the amplification of the Marcastia anticline, assuming a constant rate of bulk shortening across the structure. Any form of incremental tilting is inconsistent with simple kink-band fault-bend folding as formally described by Suppe (1983). We have attempted to model the tilt history using fixed hinge, rotating limb folding models. These are relatively successful in matching our results. Although we have only a few points on the curves they are well constrained and consistent with the modelling. We conclude that both limb rotation and minor hinge migration are likely during folding, features that are predicted from modification of buckle folds at thrust tips (Butler 1992), although existing quantitative models (e.g. Jamison 1987) of thrust propagation folds are as inapplicable as those for the classical fault-bend folds of Suppe (1983).

It is appropriate to highlight the degree of stratigraphic resolution necessary to investigate further the phenomena discussed in this paper. Where there has been sufficient preservation of syn-orogenic sediments it should be possible to investigate overall patterns of thrust activity in other examples. Chronostratigraphic control with a resolution of *c.* 1 Ma should permit recognition of the type of simultaneous thrusting patterns described in this contribution. Thus the broad timing of individual structures is generally possible.

Further constraints on possible kinematic models of folding can come from consideration of the geometry of the stratal surfaces across folds (e.g. Poblet *et al*. 1997) but only for specific, well-constrained palaeoenvironments where slopes and elevations can be reconstructed. The temporal resolution that we claim relies on a well-calibrated astro-stratigraphy. At present this does not exist for sediments older than the late Tortonian (Hilgen *et al.* 1995). There are, however, many possible sites to apply even this limited range to folds around and beyond the Mediterranean. But, it is unclear how stratigraphic data can be used to quantify tilt rates and thus test kinematic models for folding where folding rates greatly exceeded those determined here or for older or subaerial examples that do not possess a well-calibrated cyclicity.

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